Pluvial periods in Southern Arabia over the last 1.1 million-years

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Pluvial periods in Southern Arabia over the last 1.1 million-years.

Abstract

Past climates and environments experienced by the Saharo-Arabian desert belt are of prime importance for palaeoclimatic and palaeoanthropological research. On orbital timescales transformations of the desert into a savannah-like landscape in response to higher precipitation provided "windows of opportunity" for hominin dispersal from Africa into Eurasia. On long timescales, palaeoenvironmental reconstructions for the region are predominantly derived from marine sediments and available terrestrial records from the Arabian Peninsula are limited to 450 ka before present (BP). Here, we present a new stalagmite-based palaeoclimate record from Mukalla Cave in

Yemen which extends back to ~1.1 million years BP or Marine Isotope Stage (MIS) 31, as determined by Uranium-lead dating. Stalagmite Y99 grew only during peak interglacial periods and warm substages back to ~1.1 Ma. Stalagmite calcite oxygen isotope ($\delta^{18}O$) values show that every past interglacial humid period was wetter than the Holocene, a period in which large lakes formed in the now arid areas of southern Arabia. Carbon isotope ($\delta^{13}C$) values indicate habitable savannah-like environments developed during these pluvial periods. A total of 21 pluvial periods with precipitation of more than 300 mm yr^{-1} occurred since ~1.1 Ma and thus numerous opportunities for hominin dispersals occurred throughout the Pleistocene. New determinations of hydrogen (δD_{FI}) and oxygen ($\delta^{18}O_{FI}$) isotopes in stalagmite fluid inclusion water demonstrates that enhanced precipitation in Southern Arabia was brought by the African and Indian Summer Monsoons. When combined with sub-annual calcite analysis of $\delta^{18}O$ and $\delta^{13}C$, these data reveal a distinct wet (summer) and dry (winter) seasonality.

<u>Highlights</u>

- Pluvial periods recorded in stalagmites from Southern Arabia up to 1.073 Ma (MIS 31)
- Speleothem growth in Yemen only occurred during interglacial maxima and warm substages
- The African Summer Monsoon (ASM) and Indian Summer Monsoon (ISM) increased precipitation to Southwestern Arabia
- Monsoonal rainfall increased precipitation to south-eastern Arabia
- All Pleistocene pluvial periods were wetter than the Holocene pluvial period
- Grassland environments formed during peak interglacials
- Interglacial grasslands provided “windows of opportunity” for hominin occupation of the now arid Arabian interior and dispersals from Africa.

Keywords

Human dispersal, Middle East, Pleistocene, Speleothems, Arabia, Oxygen-isotopes, Carbon-isotopes, Water-isotopes, Uranium-series dating, Monsoon.

1. Introduction

The Saharo-Arabian desert belt is a key-area for both palaeoclimatic and palaeoanthropological research. On orbital timescales, changes in the intensity and spatial extent of the African (ASM) and Indian Summer monsoons (ISM) transformed the Saharo-Arabian desert belt into a “green” landscape with abundant lakes (Drake et al., 2011; Fleitmann et al., 2011; Rosenberg et al., 2011, 2012, 2013; Bretzke et al., 2013; Larrasoaña et al., 2013; Matter et al., 2015; Breeze et al., 2016; Drake and Breeze, 2016). The timing and duration of these humid periods were pivotal “windows of opportunity” for hominin dispersals from Africa into Eurasia (“out-of-Africa”), which caused substantial demographic shifts during the last 130 ka (Timmermann and Friedrich, 2016; Bae et al., 2017). Knowledge of the “permeability” of the Saharo-Arabian desert belt on longer timescales could therefore be linked to potentially earlier hominin dispersals (e.g., Hershkovitz et al., 2018; Harvati et al., 2019). To date, two dispersals routes into Eurasia are favoured, the Levantine corridor (the *northern route*) and the narrow strait of Bab-al-Mandab (the *southern route*) (Fernandes et al., 2006; Fleitmann et al., 2011; Lambeck et al., 2011; Grant et al., 2012; Rohling et al., 2013; Breeze et al., 2016).

Marine sediments from the Mediterranean (ODP 967, Larrasoana et al., 2003; Grant et al., 2017), the Red Sea (KL 11, Fleitmann, 1997) Gulf of Aden (KL 15, Fleitmann, 1997; RC09-166, Tierney et al., 2017) and Arabian Sea (ODP 721/722, deMenocal, 1995; Clemens and Prell, 2003) provide long and continuous records of climate changes in the Saharo-Arabian desert belt, with a few extending back to the Pliocene. The majority of these records use terrigenous dust as a proxy for continental wetness, where reduced dust input and grain size data are related to enhanced vegetation cover during periods of higher precipitation (Fleitmann, 1997; Larrasoana et al., 2003). However, mobilisation, transport and deposition of dust is determined by multiple non-linear factors, such as production of dust, transport paths (wind direction), wind strength, erosion and vegetation density (Zabel et al., 2001). Terrestrial archives are thus required to test and mitigate uncertainties within marine dust records.

Terrestrial records from the main dispersal routes (Fig. 1) are primarily based on lacustrine sediments and speleothems (Burns et al., 2001; Armitage et al., 2007; Vaks et al., 2010; Fleitmann et al., 2011; Petraglia et al., 2011; Rosenberg et al., 2011, 2012, 2013; Jennings et al., 2015b), which cover only the last 350 to 450 ka before present (BP) (Rosenberg et al., 2013; Parton et al., 2018). While lake records provide information on the timing of these pluvial periods, it is much more difficult to use them for characterizing the climatic conditions at the time of their formation (Rosenberg et al., 2011, 2012, 2013). Palaeolake formations currently only provide limited “wet” or “dry” environmental information; comparison of climates among interglacial periods is much more challenging. Moreover, the nature of the lakes is the subject of debate, i.e. whether seasonal "wetlands" or perennial lakes existed (Enzel et al., 2015; Engel et al., 2017; Quade et al., 2018). Furthermore, palaeolake records from Arabia cannot currently be used to determine the source of moisture; a contentious issue within palaeoclimate research (Fleitmann et al., 2003b; Rosenberg et al., 2013; Kutzbach et al., 2014; Jennings et al., 2015b; Torfstein et al., 2015). Thus, an independent archive of continental wetness is required to elucidate these issues.

Speleothems (stalagmites, stalactites and flowstones) from the Arabian Peninsula and Middle East have great potential to deliver more comprehensive climatic records as they are protected from erosion. In addition, they can be used to extend the terrestrial palaeoclimate record beyond 600 ka using the Uranium-Lead (U-Pb hereafter) chronometer (Woodhead et al., 2006, 2012; Vaks et al., 2013, 2018). In arid regions such as Arabia, speleothem growth is dependent on both availability of moisture and vegetation respired CO_2 in soils (Burns et al., 1998; McDermott, 2004). The amount and source of precipitation are important controls on speleothem calcite $\delta^{18}O_{ca}$ values (Dansgaard, 1964; Fleitmann et al., 2003a, 2011); whereas carbon isotopes ($\delta^{13}C_{ca}$) can provide information on the type (C_3/C_4 plants) and density of vegetation above the cave (McDermott, 2004; Cerling et al., 2011; Rowe et al., 2012). Finally, δD_{FI} and $\delta^{18}O_{FI}$ values of water trapped in speleothem fluid inclusion provide direct evidence of moisture sources when compared to modern isotopes in precipitation and regional meteoric waterlines (Bar-Matthews et al., 1996; Dennis et al., 2001; Meckler et al., 2015).

Previously published stalagmite records from Mukalla Cave in Yemen and Hoti Cave in Northern Oman (Fig. 1) extend back to ~330-300 ka BP, or Marine Isotope Stage (MIS) 9 (Fleitmann et al., 2011). The unique geographical position of Mukalla cave means speleothem growth occurs only when the northern limit of the monsoon rain belt passes ~14°N. Stalagmite Y99 (Mukalla Cave) is therefore an ideal specimen to track both meridional and zonal movements of the monsoon rain belt in southern Arabia and eastern Africa. Here, we present new Uranium-Thorium (^{230}Th) and Uranium-Lead (U-Pb) dates for stalagmite Y99, which allows us to expand the Arabia terrestrial palaeoclimate record back to ~1.073 Ma, or MIS 31. Additional isotope measurements performed on Mukalla and Hoti Cave stalagmite calcite and fluid inclusion water allow us to track changes in the amount and source of rainfall.

2. Climatic and Cave settings

Stalagmites presented in this study were collected from Mukalla Cave in Yemen and Hoti Cave in Northern Oman (Burns et al., 2001; Fleitmann et al., 2003b, 2011). Present-day climate in Southern Arabia is strongly governed by two major weather systems: The North Atlantic/Siberian pressure system in winter/spring and the ASM/ISM in summer (Fleitmann et al., 2003b). At present, hyper-arid to arid climate conditions prevail on the Arabian Peninsula and only the southernmost parts, such as the Yemen Highlands and Dhofar Mountains, are affected by the ASM and ISM.

2.1 Mukalla Cave, Yemen

Mukalla Cave (14°55'02''N; 48°35'23'' E; ~ 1500 metres above sea level, masl) is situated in the arid desert of Yemen, approximately 70 km North of Al Mukalla, Hadhramaut (Fig. 1). The current climate of Southern Yemen is dependent on the annual northward movement of the Intertropical

Convergence Zone (ITCZ) and associated monsoonal rainfall belt. Annual precipitation is highly variable, yet averages ~120 mm yr^{-1}, mostly delivered in the spring and summer months (Mitchell and Jones, 2005). Bedrock thickness above the cave is approximately 30 m, and soil above the cave is mostly absent. No actively growing stalagmites were found when stalagmites Y99, Y97-4 and Y97-5 were collected in 1997 and 1999 respectively (Fleitmann et al., 2011), indicating that modern rainfall is too low to recharge the aquifer above Mukalla Cave. Based on these samples, Fleitmann et al. (2011) produced an environmental record up to MIS 9 (~330 ka), identifying four distinct growth intervals (GI I-IV) within stalagmite Y99. However, only the top section (collected in whole; Fig. 2B and S1) of a 3.2m sample (Y99) was analysed. Here, remaining growth intervals from the lower part of Y99 (which was cored in several overlapping sections; Fig. 2C, S2 and S3), was dated to expand the terrestrial palaeoclimate record of Arabia. Calcite isotope measurements were performed throughout these growth intervals to characterise the climatic and environmental conditions during stalagmite growth. Additional calcite isotope measurements were performed at greater resolution in the top section of Y99.

2.2 Hoti Cave, Oman

Hoti Cave (23°05’N; 57°21’E: ~ 800 masl, Fig. 1) is located in the northern Oman mountains, where annual precipitation ranges between 50 and 255 mm yr^{-1} (station Al Hamra, 700 masl, 1974–1997). Precipitation is highly variable and mainly derived from three sources: the Mediterranean frontal system (December-March: Weyhenmeyer et al., 2002); orographic rain produced over the Jabal Akhdar Mountains during summer; and tropical cyclones, originating in the south-eastern Arabian Sea and the Bay of Bengal, every 5 to 10 years (Pedgley, 1969).

Stalagmites from Hoti Cave have been extensively studied (Burns et al., 2001; Neff et al., 2001; Fleitmann et al., 2003b, 2007). Several stalagmites cover the Holocene (samples H5, H12 and H14) and beyond (samples H1, H4, and H13). Stalagmite H13 is a ~3 m tall stalagmite covering MIS 5e, MIS 7e and MIS 9. Further details on the chronology and sampling location of Hoti Cave were presented in previous publications (Burns et al., 1998, 2001; Neff et al., 2001; Fleitmann et al., 2003b, 2007, 2011).

3 Methods

3.1 Dating

Stalagmites presented in this study were dated using the ^{230}Th dating method back to ~550 ka and the U-Pb method for older samples (Woodhead et al., 2006; Cheng et al., 2013). The ^{230}Th ages for Hoti Cave stalagmites are reported in Fleitmann et al. (2007, 2003a). For stalagmite Y99 (Mukalla Cave), a total of seventy ^{230}Th ages were determined back to approx. 550 ka BP (Tab. S1-S3). Nineteen samples were analysed at the University of Minnesota (following the methods outlined by Cheng et al., 2013) and twelve additional samples were analysed at the British Geological Survey, Nottingham, UK (following the methods outlined by Crémière et al., 2016). Dates were calculated using the decay constant of Cheng et al. (2013), and a correction for the presence of initial ^{230}Th was applied assuming a detrital U-Th isotope composition of (^{232}Th/^{238}U) = 1.2 ± 0.6, (^{230}Th/^{238}U) = 1 ± 0.5 and (^{234}U/^{238}U) = 1 ± 0.5. The ages for GI XII and GI XVIII were determined via U-Pb methods. U-Pb ages for the lower part of Y99 were produced using both traditional solution-mode multi-collector inductively coupled plasma mass spectrometry (MC-ICP-MS) (following the methods detailed in: Woodhead et al., 2006) analysis (University of Melbourne, Australia) along with the recently developed Laser ablation (LA) method (BGS) (Tab. S4 and S5). For LA-ICP-MS, the methods and analytical protocol follows that described by Coogan et al. (2016); U/Pb ratios were normalised to WC-1 carbonate (Roberts et al., 2017) and Duff Brown carbonate (Hill et al., 2016) was run as a check on accuracy.

3.2 Calcite oxygen and carbon isotope analysis

A total of 910 samples were collected along the main growth axes of stalagmite Y99 GIs for stable isotope analysis. Samples were collected at resolutions of ~1mm for growth phase I and II, ~2 mm for growth phases III-VII and ~5 mm for growth phases VII-XVIII (Tab. S6). Due to the variable size, visibility and direction of independent growth layers, it was not possible to produce Hendy tests. To provide addition support for our Growth Interval assignments, additional samples were collected across visible growth discontinuities at 1 mm resolution within the lower sections of Y99 (Tab. S7; Fig. 2). Furthermore, H13 (Hoti Cave) was selected for sub-annual isotopic analysis to examine seasonality, due to its annual laminations. Samples were collected at 0.1mm resolution (Tab. S8).

Isotope measurements were performed using a Finnigan Delta V Advantage Isotope Mass Spectrometer (IRMS) coupled to an automated carbonate preparation system (Gasbench II). Precision (1σ) is ≤0.2‰ for $\delta^{18}O$ and ≤0.1‰ for $\delta^{13}C$. Measurements were performed at the Chemical Analysis Facility (CAF), University of Reading, UK, and the Institute of Geological Sciences, University of Bern, Switzerland. Isotope values are reported relative to the Vienna Peedee Belemnite (VPDB) standard.

3.3 Fluid inclusion deuterium and oxygen isotope analysis

Deuterium (δD_{FI}) and oxygen ($\delta^{18}O_{FI}$) isotopes of speleothem fluid inclusion water were analysed at the Physics Institute, University of Bern, Switzerland, using a recently developed extraction method (Affolter et al., 2014, 2015). Sixteen calcite blocks of ~25 x 5 x 5 mm (L, W, H) for fluid inclusion analysis were collected from Y99, H13 and H5. Samples were placed into a copper tube and connected to the measuring line, heated to ~140°C and crushed, the liberated water was then transported to a wavelength scanned cavity ring down spectroscopy system (Picarro L2401-i analyser) under humid conditions (with standardised water of known isotopic composition) to prevent fractionation and minimize memory effects. The crushing of samples released, on average, ~1 μl of water. Precision is 1‰ for δD_{FI} and 0.2‰ for $\delta^{18}O_{FI}$. Fluid inclusion values are reported on the Vienna Standard Mean Ocean Water (V-SMOW) scale (Tab. S9).

4. Results and Discussion

This section is divided into two parts. In the first part we focus on rainfall variability during the last 350 ka. We provide additional and more precise ^{230}Th ages for Y99 (Mukalla Cave), as well as stable isotope analysis of calcite and fluid inclusion water from Y99 (Mukalla Cave), H5 and H13 (Hoti Cave). We combine these ages with previously published Mukalla and Hoti Cave speleothem data to discuss the timing and environmental conditions of South Arabian Humid Periods (SAHPs) since 350 ka BP. By comparing our multiproxy records with marine and terrestrial palaeoclimate records from the African and Asian monsoon domains, we show that periods of enhanced rainfall and speleothem growth in Southern Arabia are related to a strengthening and greater spatial extent of the ASM and ISM during peak interglacials and interstadials. Within the second section, we provide an extended chronology and $\delta^{18}O_{ca}$ and $\delta^{13}C_{ca}$ stable isotope data for the lower portion of Y99 (Fig. 2C) in order to characterise humid periods in Southern Arabia back to ~1.1 Ma BP, making the stalagmite Y99 record one of the longest continental records from Southern Arabia.

4.1 Timing and Nature of SAHPs during the last 350 ka

4.1.1 Chronology of Y99 GI I to V

The chronology of Y99 growth phase I to V is based on a total of 53 ^{230}Th ages (Fig. 3). These include 38 ages presented in Fleitmann et al. (2011) and 15 additional more precise ^{230}Th ages analysed for this study (Tab. S1 and S2). Stalagmite Y99 GIs I-V coincide with peak interglacial periods and interstadials corresponding to MIS 5e, 7a, 7e, 9c and 9e (Fig. 4) when Southern Arabia was affected by the ASM and ISM (Fleitmann et al., 2011; Rosenberg et al., 2013). While age reversals are observed in GI IV and V, kernel probability density plots of all ^{230}Th ages obtained from Mukalla Cave (Y99, Y97-4.-5) and Hoti Cave (H1, H4, H5, H10, H11 and H14) indicates Y99 growth was more likely to occur within MIS 9c and 9e (Fig. 3).

4.1.2 SAHPs in Oman and Yemen during the last 350 ka

Growth intervals of stalagmites from Mukalla and Hoti Caves mark climatic intervals when effective precipitation was high enough to recharge the aquifers above both caves (Burns et al., 2001; Fleitmann et al., 2003b, 2011; Fleitmann and Matter, 2009). At present, total annual rainfall averages ~120 mm

yr^{-1} and ~180 mm yr^{-1} at Mukalla and Hoti Cave respectively, and actively growing stalagmites are either absent (Mukalla Cave) or very rare (Hoti Cave) (Burns et al., 2001; Fleitmann et al., 2003a, 2007, 2011). Thus, the existence of very tall and large diameter stalagmites such as H13 and Y99 (Fig. 2) in both caves is clear evidence that precipitation was considerably higher than today when they were formed (Vaks et al., 2010; Fleitmann et al., 2011; El-Shenawy et al., 2018). Based on the spatial distribution of actively growing stalagmites in the Levant and Negev – areas with very similar climatic conditions compared to Yemen and Oman – precipitation should have been around 300 mm yr^{-1} or greater to recharge the groundwater and trigger growth of stalagmites (Vaks et al., 2010, 2013). Considering the height and diameter of stalagmites Y99 and H13 (Fig. 2), precipitation was most likely considerably higher than 300 mm yr^{-1}. Intervals of speleothem growth at both cave sites are therefore a first indicator for continental wetness in Southern Arabia. An important feature of stalagmites Y99 and H13 is that their growth was reactivated multiple times, suggesting that the long-lasting cessations of stalagmite growth are related to arid climatic conditions (Burns et al., 2001; Fleitmann et al., 2011).

Over the last 350 ka BP, stalagmite growth in Mukalla and Hoti Caves (Fig. 4) occurred during peak interglacial periods and warmer substages corresponding to the early and mid-Holocene, MISs 5a, 5c, 5e, 7a, 7e, 9c and 9e (Fig. 4; Burns et al., 2001; Fleitmann et al., 2003a, 2003b; 2011; Fleitmann and Matter, 2009). SAHPs were related to intensified African and Indian summer monsoon circulation and a northward displacement of the tropical rain-belt and ITCZ at times of high boreal summer insolation and low ice volume (LR04 stack) (Burns et al., 1998, 2001; Fleitmann et al., 2011; Beck et al., 2018). Both the timing and frequency of SHAPs over the last 350 ka are in excellent agreement with other marine and terrestrial hydroclimate records from the Saharo-Arabian desert belt (Fig. 4). In the Gulf of Aden, low $\delta D_{leafwax}$ values in Core RC09-166 (Fig. 4) indicate greater rainfall in the Horn of Africa and Afar regions during the early to mid-Holocene (SAHP 1), MIS 5a, 5c, 5e (SAHPs 2-4) and MIS 7a (SAHP 5) (Tierney et al., 2017). Two aeolian dust records from the Gulf of Aden (KL 15) and central Red Sea (KL 11) show generally lower median grain size values during peak interglacial periods when erosion and mobilization of dust was significantly reduced as a result of a denser vegetation cover in North

Africa and Arabia (Fleitmann, 1997). Similarly, speleothem growth in Southern Arabia and Northern Egypt (Wadi Sannur Cave) are in good agreement, occurring at MIS 5e, MIS 7c and MIS 9c and 9e (El-Shenawy et al., 2018). Absence of speleothem growth in Northern Egypt during relatively warm substages (MIS 5c and 5a and MIS 3) suggests that ASM rainfall did not reach far into Egypt, highlighting a degree of regional and temporal variability. Sapropel layers in the Eastern Mediterranean are an additional proxy for ASM and ISM intensity and were mainly deposited during periods of increased Mediterranean rainfall and significantly higher monsoon precipitation in the Ethiopian highlands and resultant higher Nile discharge (Fig. 4; summarized in Rohling et al., 2015; Grant et al. 2016). The timing of SAHPs 1 to 8 is in excellent agreement with sapropels records, with the exception of the “ghost sapropels” 2 and 6 which are most likely not associated with higher Nile discharge (Rohling et al., 2015). Further north, the Soreq and Peqiin Cave $\delta^{18}O_{ca}$ records from the Levant are also sensitive recorders of changes in $\delta^{18}O$ of eastern Mediterranean surface seawater related to Nile discharge (Bar-Matthews et al., 2003; Rohling et al., 2015), with more negative $\delta^{18}O_{ca}$ values indicating higher Nile discharge during peak interglacial and interstadial periods (Fig. 4). Likewise, speleothem-based Negev Humid Periods (NHPs; based on speleothem ages) 1-4 are synchronous to SAHPs (Fig. 4), with the exemption of SAHP 6 (~245-241 ka), in which there is only limited evidence of speleothem deposition (Vaks et al., 2010). Also, SAHPs 1-8 correlate to phases of lake formation in the Nafud desert in Northern Arabia related to enhanced ASM rainfall (Rosenberg et al., 2013; Jennings et al., 2015b). SAHPs are therefore in phase with wet intervals in Northern Arabia. One notable discrepancy, however, is the lack of evidence for stalagmite growth in Mukalla and Hoti Caves during MIS 7c (Fig. 4); whereas increased precipitation is observed in Wadi-Sannur Cave, Peqiin and Soreq $\delta^{18}O_{ca}$ records, KL-15 grain size and Mediterranean sapropels (S8) (Fig. 4). MIS 7c is also reflected by a less substantial enhancement of the monsoon in Asia (Beck et al., 2018) and KL-11 (Fleitmann et al., 1997) (Fig. 4). The reason of lack of evidence for an SAHP during MIS 7c remains unknown. Furthermore, we acknowledge that some fluvio-lacustrine deposition and alluvial aggradation occurred in Arabia during MIS 6 and 3 (e.g., McLaren et al., 2009; Parton et al., 2013, 2015, 2018; Hoffmann et al., 2015). Previous analyses have shown that that only 200 mm yr^{-1} is required to activate alluvial systems in Arabia (Parton et al., 2015); whereas more than 300 mm yr^{-1} required to active the growth of tall stalagmites (Vaks et al., 2010; Fleitmann et al., 2011).

The influence of precessional and glacial boundary forcing on Asian monsoon intensity remains controversial, as some monsoon records suggest dominant precession-driven monsoon maxima during Northern hemisphere summer insolation maxima (Cheng et al., 2016) while others show evidence for a dampening effect of glacial boundary conditions on monsoon strength during glacial periods (Burns et al., 2001; Fleitmann et al., 2003a; Beck et al., 2018). A recently published East Asian summer monsoon (EASM) reconstruction based on ^{10}Be-flux from Chinese loess shows highest summer monsoon rainfall during peak interglacial periods (Fig. 4). The ^{10}Be-flux rainfall EASM record is closely linked with global ice volume, which is consistent with the timing of SAHPs 1-8 in our speleothem record (Fig. 4).

In summary, there is excellent agreement between SAHPs and low latitude northern-hemisphere insolation, glacial boundary conditions and African and Asian (Indian) monsoon records. This adds confidence that the Mukalla and Hoti Cave speleothems are an accurate recorder of changes in ASM and ISM intensity and extent in north-eastern Africa and Southern Arabian Peninsula.

4.1.3 Source of moisture in Southern Arabia during SAHPs 1 to 8

Current climate reconstructions derived from lacustrine sediments and dune deposits are unable to identify the source of precipitation during Arabian pluvial periods (Fleitmann et al., 2003b; Kutzbach et al., 2014; Enzel et al., 2015; Torfstein et al., 2015; Engel et al., 2017). This has triggered controversial debates about the origin of rainfall at the time of their formations. Enzel et al. (2015), for instance, questioned the paradigm that enhanced precipitation in Arabia was related to an amplification of the ASM and ISM and northward displacement of the summer ITCZ. Instead, Enzel et al. (2015) proposed two other potential sources of precipitation in Oman during the early and mid-Holocene humid period (SAHP 1): more frequent Arabian Sea cyclones or enhanced advection of moisture from the Gulf of

Oman in winter. Direct measurements of hydrogen and oxygen isotope values in speleothem fluid inclusions from Hoti and Mukalla Caves provide direct information on drip water isotopic composition and palaeoprecipitation respectively (Fleitmann et al., 2003b). Stalagmite δD_{FI} and $\delta^{18}O_{FI}$ values enable us to determine the origin (e.g., Mediterranean or Indian Ocean) and transport of moisture to Southern Arabia during pluvial periods. Furthermore, they also permit a direct comparison with isotope-enabled climate model simulations, to benchmark the models (Herold and Lohmann, 2009) and also help to settle current debates about the origin and seasonality of precipitation in Southern Arabia.

At present, a large proportion of moisture in Yemen derives from the northern reach of the ISM (Fleitmann et al., 2011) with additional moisture originating from Africa (by the ASM) and the Red Sea (mainly in winter/spring) (e.g., Al-ameri et al., 2014). The isotopic composition of modern precipitation (collected between 2008 and 2010) from sampling sites between 500 and 1700 meters asl in Yemen ranges from around -40 to 40‰ and -4 to 8‰ in δD and $\delta^{18}O$ respectively, and rainfall plots along the Global Meteoric Waterline (GMWL; δD=8 $\delta^{18}O$ + 10; Fig. 5A; Al-ameri et al., 2014). In contrast, stalagmite Y99 fluid inclusion isotope values for MISs 5e (SAHP 4) and 7e (SAHP 6) are more negative and range from -64.5‰ to -35.0% and -8.6 and -4.5‰ in δD and $\delta^{18}O$ respectively (Fig. 5A; Tab. S9). Like modern rainfall, Y99 fluid inclusion isotope values plot close to the GMWL, whereas some samples appear to be slightly affected by evaporation as they plot below the GMWL. Stalagmite Y99 MIS 5e and MIS 7e δD_{FI} and $\delta^{18}O_{FI}$ values are more negative than isotope values in modern summer monsoonal rainfall (June to September) in Addis Ababa, Ethiopia, where moisture is delivered by the African and Indian summer monsoons. This suggests that enhanced rainfall at Mukalla Cave during MIS 5e (SAHP 4) and MIS 7e (SAHP 6) resulted from an amplification of the ASM and/or ISM (Fig. 5C). Our assumption is also supported by climate model data for MIS 5e (Herold and Lohmann, 2009; Jennings et al., 2015b), which indicate a more zonal transport of moisture from Africa to the Arabian Peninsula during MIS 5e (SAHP 4). Y99 $\delta^{18}O_{FI}$ values of around -7.2 ± 1.5‰ during MIS 5e are within the range of modelled summer precipitation $\delta^{18}O$ values of between -6 and -7‰ in Yemen (Fig. 5C).

Finally, significant contributions of rainfall from a Mediterranean source can be excluded as Y99 δD_{FI} and $\delta^{18}O_{FI}$ values plot below the Mediterranean Meteoric Waterline (MMWL; $\delta D = \delta^{18}O + 22$; Fig. 5A).

In Northern Oman, present-day rainfall originates predominantly from a northern (Mediterranean) and a southern (Indian Ocean) moisture source. As a result, two distinctly different local meteoric waterlines exist, the Northern Meteoric Waterline (N-LMWL; $\delta D = 5.0\ \delta^{18}O +10.7$) and the Southern Local Meteoric Waterline (S-LMWL; $\delta D = 7.1\ \delta^{18}O -1.1$) (Fig. 5B; Weyhenmeyer et al., 2002; Fleitmann et al., 2003b). Precipitation originating from a northern moisture source ranges from -4.5 to 1.0‰ in $\delta^{18}O$ and from -25 to 5‰ in δD, whereas precipitation from a southern moisture source is more negative, with $\delta^{18}O$ values varying from -10 to -2‰ and δD values from -75 to -15‰ (Weyhenmeyer et al., 2002; Fleitmann et al., 2003b). Modern groundwater in Northern Oman (N-OGL: $\delta D = 5.3\ \delta^{18}O + 2.7$) and cave drip water in Hoti Cave is intermediate between both sources, indicating that both contribute to groundwater recharge (Weyhenmeyer et al., 2002; Fleitmann et al., 2003b; Fig. 5B). The isotopic composition of fluid inclusion water extracted from the Holocene stalagmite H5 (SAHP 1: 10.9 ka-6.2 ka; Neff et al., 2001; Fleitmann et al., 2007) ranges from -21.4‰ to -13.2‰ in δD_{FI}, and -3.2‰ to -0.7‰ in $\delta^{18}O_{FI}$. Fluid inclusion water extracted from the MIS 5e (SAHP 4) section of stalagmite H13 is more negative and measured -41.7‰ for δD_{FI} and -7.8‰ to -4.2‰ for $\delta^{18}O_{FI}$. Both H5 and H13 fluid inclusion values plot closer to the S-OMWL (Fig. 5B), indicating the ISM was the primary moisture source in Oman during peak interglacials (Fleitmann et al., 2003b). One sample, however, plots above the S-OMWL, yet this remains more closely aligned to modern southern groundwater values. Overall, Hoti Cave δD_{FI} and $\delta^{18}O_{FI}$ values show clear evidence that enhanced rainfall during the early to middle Holocene (SAHP 1) and MIS 5e (SAHP 4) was related to an intensification of the ISM. This is in stark contrast to suggestions that enhanced frontal depressions from the Persian Gulf and/or Mediterranean increased precipitation during the early to mid-Holocene wet period (SAHP 1) (Enzel et al., 2015).

When compared to isotope-enabled climate model simulation, the measured isotope $\delta^{18}O_{FI}$ values at both caves for SAHP 4 (MIS 5e) are in good agreement with modelled $\delta^{18}O$ of summer precipitation (Fig. 5C). Furthermore, the distinct isotopic gradient across Southern Arabia is also supported by the Y99 and H13 $\delta^{18}O_{FI}$ values, with more negative modelled summer rainfall $\delta^{18}O$ values prevailing in the west due to a greater moisture supply from the African summer monsoon (Herold and Lohmann, 2009). Thus, growth intervals and isotope values in stalagmites from Mukalla Cave are excellent proxies for the intensity of the ASM in eastern Africa, whereas stalagmites from Hoti Cave are more closely connected to intensity changes of the ISM and, to a lesser extent, ASM. This is also in agreement with climate model simulations for MIS 5e, which indicate that higher precipitation in Northern Oman was associated with the ASM and ISM, with negligible and fairly stable contribution of rainfall from Mediterranean westerlies (Jennings et al., 2015b).

4.1.4 Comparison between pluvial conditions in Southern Arabia during the last 350 ka

$\delta^{18}O_{ca}$ values of stalagmites from Southern Arabia are primarily controlled by two effects: i.e. the amount and source of rainfall (Fleitmann et al., 2003b, 2004, 2007, 2011). $\delta^{18}O_{ca}$ values of Mukalla Cave stalagmites are generally more negative than those of stalagmites from Hoti Cave (Fig. 6A), with a west-east (Mukalla-Hoti) isotopic gradient of between 2 and 4 ‰ during SAHPs 1-7. This gradient is also evident in δD_{FI} and $\delta^{18}O_{FI}$ values from both caves and in simulated MIS 5e $\delta^{18}O$ in summer precipitation across Southern Arabia (Fig. 6A). This adds further confidence in the palaeoclimatic significance of $\delta^{18}O_{ca}$ values from Mukalla and Hoti Caves. Furthermore, $\delta^{18}O_{ca}$ values from both cave sites reveals marked and consistent differences in the amount of rainfall between among the SAHPs (Fig. 8). The most striking feature of the Mukalla and Hoti Cave records is the fact that least negative $\delta^{18}O_{ca}$ values were obtained from early to mid-Holocene stalagmites, indicating that monsoon rainfall during SAHP 1 was the lowest in the last 350 ka. On the other hand, monsoon precipitation was highest at both caves during SAHP 4 (MIS 5e). When we combine our fluid inclusion data with the relatively consistent $\delta^{18}O_{ca}$ between SAHPs, we can show that the moisture source was likely consistent throughout SAHPs. Modern $\delta^{18}O_{ca}$ values from Hoti Cave (derived from the winter Mediterranean precipitation source) are more positive than SAHP values (Fig. 6A). SAHP 1 (early to middle Holocene), 4 (MIS 5e) and 6 (MIS 7e) FI data allows us to confidently state that $\delta^{18}O_{ca}$ of these periods represents a monsoon rainfall signature. Thus, we can use the more positive $\delta^{18}O_{ca}$ values (Mediterranean signature) of modern and more negative $\delta^{18}O_{ca}$ values (monsoon signature) of past precipitation to posit that monsoon precipitation was the dominant source of preceding SAHPs. This isotopic relationship has also been observed in previously published high-resolution $\delta^{18}O_{ca}$ profiles of H5 and H12 (Fig. 6B), where an abrupt shift from more negative values (increased precipitation from the ISM) to more positive values (reduced precipitation delivered by Winter Mediterranean Cyclones (WMCs)) occurred at the termination of the early Holocene pluvial period (SAHP 1) (Fleitmann et al., 2007).

4.1.5 Environmental conditions

Mukalla Cave speleothem $\delta^{13}C_{ca}$ values vary between -8 and 2‰ (VPDB) (Fig. 7; Tab. S11). Such a wide range in $\delta^{13}C_{ca}$ is quite common in speleothems as $\delta^{13}C_{ca}$ depends on a variety of environmental, partly counteracting, parameters, including: (1) type and density of vegetation, (2) soil thickness and moisture, (3) biological activity within the soil, (4) recharge conditions and (5) kinetic isotope fractionation during calcite precipitation, the latter factor is influenced by cave air PCO_2 and drip rate (e.g., Baker et al., 1997). At times of high precipitation and short soil-water residence times, equilibration between soil CO_2 and percolating water may be incomplete. Under such a scenario, seepage water would have a stronger atmospheric CO_2 component and thus speleothem $\delta^{13}C_{ca}$ values would be more positive. In addition, CO_2 degassing within the cave can lead to more positive speleothem $\delta^{13}C_{ca}$ values and thus blur the biogenic signal. Overall, speleothem $\delta^{13}C_{ca}$ values can be difficult to interpret, which is one reason why the Hoti Cave $\delta^{13}C_{ca}$ records were never used for palaeoenvironmental reconstructions. This is also related to the fact that Hoti Cave has two entrances and therefore strong ventilation, leading to fluctuations in cave air PCO_2 and strong kinetic fractionation of $\delta^{13}C$ during calcite precipitation. In contrast, Mukalla Cave has only one narrow entrance and ventilation within the cave is therefore low. Mukalla Cave stalagmite $\delta^{13}C_{ca}$ values are therefore more closely related to surface vegetation and biological soil activity, provided that complete equilibration (“open system conditions”) between soil CO_2 and soil water has occurred. Under such conditions, $\delta^{13}C_{ca}$ values of a stalagmite growing under a C_3 plant dominated environment vary between -14 and -6‰ (VPDB) and -6 to +2‰ under C_4 plants (Clark and Fritz, 1997; McDermott, 2004). Assuming open system conditions, Mukalla Cave speleothem $\delta^{13}C_{ca}$ values fall into the range of C_4 plant dominated vegetation with occasional C_3 plants (Fig. 7), indicating herbaceous semi-desert grassland environment above the cave during SAHPs 1-8.

Our data also shows that C_4 environments were present during the warm substages of MIS 5. This is in good coherence with phytolith data from the Jabal Faya archaeological site, UAE, showing denser and more diverse vegetation was present during MIS 5 than succeeding substages (Bretzke et al., 2013). Grassland taxa (*Kobus*, *Hippopotamus*, *Pelovoris*) and *H. sapiens* were uncovered from MIS 5a palaeolake sediments in the Nafud showing that grasslands were present in northern Arabia (Groucutt et al., 2018). In particularly, *Hippopotamus* is not a long-distance migratory species, and requires year-round access to water. Similarly, the MIS 5e speleothem $\delta^{13}C_{ca}$ values from Ashalim Cave, Negev, range from -8‰ to -2‰ (Vaks et al., 2010), suggesting comparable environments existed in the northern and southern extent of the Saharo-Arabian desert. Archaeological and fossils finds have demonstrated that *H. sapiens* were present in Arabia during MIS 5 interstadials (Groucutt et al., 2018). Furthermore, Mukalla Cave $\delta^{13}C_{ca}$ values are coherent with palaeontological evidences from older pluvial periods. Faunal assemblages from the Ti's al Ghadah palaeolake (MIS 9-13) exhibit large mammals from African and European sources (Thomas et al., 1998; Rosenberg et al., 2013; Stimpson et al., 2016), showing these wet periods were sufficient to sustain fauna that required a perennial water supply. Overall, our data adds to the growing evidence that the formation of widespread ‘green’ environments formed across Arabia during peak interglacial periods, which facilitated *H. sapiens* occupation and movement across the now desert areas of Arabia.

4.1.6 Seasonality of precipitation during SAHPs

Some stalagmites from Hoti Cave exhibit distinct annual layers, with a thickness varying between 0.1 and 1.2 mm (e.g., stalagmite H14; Cheng et al., 2009). Such layers are also visible un the MIS 5e section of stalagmite H13, composed of a white porous laminae and dense translucent laminae. Their presence suggests distinct seasonal changes in the drip rate in response to surface precipitation. Nearly monthly resolved $\delta^{18}O_{ca}$ and $\delta^{13}C_{ca}$ profiles over 4 years (Fig. 8B; Tab. S8) show seasonal variations of more than 1 ‰, where denser layers display more negative $\delta^{18}O_{ca}$ and $\delta^{13}C_{ca}$ values. We suggest that these denser layers were formed during the monsoon seasons, at times of higher drip rate, slower CO_2 degassing and lower evaporation of cave drip waters (Fleitmann et al., 2004). In. contrast, the more porous white layers display more positive $\delta^{18}O_{ca}$ and $\delta^{13}C_{ca}$ due to a reduced drip rate, resulting in greater CO_2 degassing and evaporation. Combined with δD_{FI} and $\delta^{18}O_{FI}$ values from H13, the presence of annual layers during SAHP 1 (early to mid-Holocene; Cheng et al., 2009) and SAHP 4 (MIS 5e; this study) and seasonal changes in $\delta^{18}O_{ca}$ and $\delta^{13}C_{ca}$ indicates that Southern Arabia experienced a rainy (monsoon) season during boreal summer and a drier season during boreal winter. This is in good agreement with climate simulations for MIS 5e, with simulations at 130, 125 and 120 ka BP (Gierz et al., 2017). These simulations show a strong increase in summer (JJA) precipitation at 130 and 125 ka BP, whereas no significant increase in winter (DJF) is observed in Southern Arabia (Fig. 8C). We can therefore exclude that increased precipitation was provided by enhanced Mediterranean cyclone activity in winter/spring as suggested Enzel et al. (2015). Taken together, there is clear evidence that climatic conditions during SAHPs were still characterized by a strong seasonality with wet summers and rather dry winters.

<u>4.2 Timing and Nature of SAHPs beyond 350 ka</u>

4.2.1 Chronology of stalagmite Y99 beyond 350 ka

The identification of Y99 GIs beyond 350 ka BP is based (1) on thirty-one ^{230}Th and three U-Pb ages (Tab. S1, S2 and S4), (2) macroscopic evidence for major discontinuities (e.g., abrupt changes in colour of the fabric, abrupt changes in the frequency of laminae, finely defined and bright laminae and lateral displacements of the growth axis), and (3) abrupt shifts in $\delta^{18}O_{ca}$ over potential discontinuities (Fig. 2D). The latter are strong evidences for the termination of a SAHP, as they occur immediately before the cessation of stalagmite growth, when annual precipitation dropped below 300 mm yr^{-1} (Burns et al., 2001; Fleitmann et al., 2003b). Positive shifts in $\delta^{18}O_{ca}$ are also observed in stalagmites from Hoti Cave, where they mark a weakening and termination of pluvial monsoon periods within a few decades. This is particularly evident ~6.2 ka BP (Neff et al., 2001; Fleitmann et al., 2007; Fig. 6B). Using these criteria, we identified 12 further GIs (VI to XVIII) in stalagmite Y99. However, it is slightly more challenging to assign absolute ages to these GIs, as it becomes increasingly difficult to obtain accurate and precise ages ^{230}Th ages beyond 400 ka BP. While the chronology of Y99 GIs VI-XVIII is supported by thirty-one ^{230}Th and three U-Pb ages, ^{230}Th ages for GIs VI to VIII are not in perfect stratigraphic order, with several age reversals occurring between 400 and 550 ka BP. This is not related to analytical problems but rather caused by post-depositional mobilisation of U and Th, potential small-scale dissolution and re-precipitation of calcite or incorporation of ^{230}Th adsorbed to organic acids (Borsato et al., 2003; Scholz et al., 2014). These effects can imply localized open-system behaviour (Bajo et al, 2016). While post-depositional leaching of U would lead to older ages, re-precipitation of calcite or incorporation of ^{230}Th would result in younger ages as observed in some GIs. All these effects are critical for very old samples that are close to the ^{230}Th-dating limit of ~500-600 ka as even minute post-depositional alterations and several phases of dissolution and/or re-precipitation can have significant effects on the age. The higher porosity and micro-voids make in the upper section of stalagmite Y99 (Fig. 2A and B) more prone to post-depositional loss or addition of radionuclides. In contrast, the lower part of Y99, comprising GIs IX to XVIII, is composed of very dense calcite but too old for the ^{230}Th-dating method. Nevertheless, two U-Pb ages determined in different laboratories are consistent and date the base of stalagmite Y99 to 1.07 ± 0.04 Ma (GI XVIII; MIS 31). One additional U-Pb age of 0.85 ± 0.07 Ma BP for GI XII serves as an additional tie point for the chronology of the lower part of stalagmite Y99. Based on the consistent pattern of high-monsoonal rainfall and stalagmite growth during interglacial intervals during the last 350 ka BP (Fig. 4), we used orbital tuning to the LR04 stack (Lisiecki and Raymo, 2005) to assign absolute ages for Y99 GIs VI to XIII and XIV to XVII (Fig. 9). The good match between the number of identified GIs and peak interglacial periods gives credence to the Y99 chronology.

4.2.2 Climate and environmental conditions in Southern Arabia over the last 1.1 Ma

At least 21 SAHPs occurred over the last 1.1 Ma at times when low ice volume and high summer insolation strengthened both the ISM and ASM. AS mentioned previously, there is a general scarcity of terrestrial records covering more than 400 to 500 ka BP in Northern Africa and the Arabian Peninsula, and marine sediments are the only source of information. Two dust records from the Eastern Mediterranean (ODP 967; Grant et al., 2017) and Arabian Sea (ODP 721/722; deMenocal et al., 1995) extend beyond 500 ka BP and are interpreted to reflect continental wetness in the wider northeast African region and the Arabian Peninsula (Fig. 10). The ODP 967 PCA index of rainfall and aridity (PC2; Fig. 10) is based on trace element content (e.g., titanium) and sapropels and shows distinct fluctuations in the strength and spatial extent of the ASM. The ODP 967 record thus provides evidence for multiple “Green Sahara Periods”. When compared to the stalagmite record of SAHPs, some wet periods in the ODP 967 record coincide with SAHPs, such as during the early to middle Holocene, MIS 5e, 7a, 9e and 13a (Fig. 10). There are also notable differences and wet climatic phases in the ODP 967 PC2 record are not always consistent with SAHPs, such as MIS 6 or MIS 16. Adversely, between 950-650 ka BP, the ODP 967 rainfall index is typically in a ‘dry’ mode while at least three SAHPs occurred within MIS 17, 19 and 21. Likewise, the association between SAHPs and low dust content in the ODP 721/722 core is not always evident. For instance, dust content is relatively low and constant between MISs 12 and 16, suggesting rather humid climatic conditions between 425 and 675 ka BP. The discrepancy between dust and stalagmite records has been observed before (Fleitmann et al., 2011) and could be related to regional variability, availability of dust and changes in wind direction and strength.

Stalagmite Y99 $\delta^{18}O_{ca}$ values of all GIs are very similar and typically range from -7 to -11 ‰ (Fig. 9), indicating that the ASM was the dominant source of precipitation during all SAHPs. There are, however, differences in the degree of wetness between SAHPs as more negative $\delta^{18}O_{ca}$ values indicate higher ASM and ISM rainfall in Yemen and Oman. The boxplot (Fig. 9) of $\delta^{18}O_{ca}$ values of all Mukalla Cave stalagmites show that SAHPs 4, 5, 18, 19 and 20 exhibit the most negative $\delta^{18}O_{ca}$ values and are thus characterized by the highest monsoonal rainfall.

Y99 $\delta^{13}C_{ca}$ values range 2 to -8‰ (Fig. 9) and differences are apparent between SAHPs. As stated above, these ranges are typical of C_4 environments above the cave assuming 'open system' conditions. However, $\delta^{13}C_{ca}$ values can be influenced by various parameters such as vegetation type and density, soil thickness and moisture, as well as atmospheric and other processes (McDermott, 2004; Rowe et al., 2012). Moreover, deluge of thin soils at times of very high rainfall can lead to more positive speleothem $\delta^{13}C_{ca}$ values due to rapid infiltration into the karst system and reduced interaction with soil CO_2 (e.g., Bar-Matthews et al., 2003). Increased rainfall during SAHP 19-20 could have led to more positive $\delta^{13}C_{ca}$ values. In contrast, reduced rainfall and increased interaction of percolating water with soil CO_2 may have had the opposite effect, contributing to more negative $\delta^{13}C_{ca}$ values during SAHP 14-17. Due to the numerous controls on speleothem $\delta^{13}C_{ca}$, alterations of the principal determinant can be expected over such a long period of time. Despite this, the overall range of $\delta^{13}C_{ca}$ values indicates C_4 grasslands were present during SAHP VI-XVII. This shows that interglacial periods routinely saw vegetation form in the now desert areas of southwestern Arabia.

4.3 Hominin migrations

4.3.1 Early-Middle Pleistocene

Estimates for the potential timing of hominin dispersals during the last few hundred thousand years are mostly modelled on palaeoclimate conditions of East Africa (deMenocal, 1995; Shultz and Maslin, 2013; Maslin et al., 2014) or Eurasia (Muttoni et al., 2010; Kahlke et al., 2011). These models do not consider whether and when the Saharo-Arabian desert was traversable. Yet the formation of so called "green corridors" between sub-Saharan Africa, northern Africa and Eurasia created "windows of opportunity" that would have been critical for hominin occupation and dispersal. Though, it is surely more apt to consider these areas as "green landscapes" in which hominin populations inhabited – rather than a route to the 'other' side. Based on the timing of SAHPs and their close connection to humid intervals in Northern Africa, we suggest that the Saharo-Arabian grasslands could facilitate occupation and dispersal during MIS 31 (~1080 ka: SAHP 21), MIS 29 (~1014 ka: SAHP 20), MIS 28b (~1000 ka: SAHP 19) MIS 27a (~982 ka: SAHP 18) MIS 25 (~955 ka: SAHP 17), MIS 21 (~850 ka: SAHP 16) and MIS 19 (~760 ka: SAHP 15) (Fig. 10). Frequent windows of opportunity take place between SAHP 21 to SAHP 17 (MIS 31 to MIS 25), varying between 40-10 ka intervals. SAHP 21 to 18 are also marked as some of the most negative $\delta^{18}O_{ca}$ values in Y99 (Fig. 8), indicating intense monsoon periods. Succeeding this, SAHPs reduced in frequency, with ~100-70 ka intervals between SAHPs 17-13 (MIS 25-15e). These are also marked by increased $\delta^{18}O_{ca}$ values – indicating somewhat drier periods – and reduced occurrence of Mediterranean sapropels (Fig. 10). This shift echoes the transition from ~40 ka to ~100 ka glacial interglacial cycles, known as the Middle Pleistocene Transition (MPT) (Lisiecki and Raymo, 2005; Railsback et al., 2015; Tzedakis et al., 2017). Thus, it is likely that the pattern of hominin occupation of Arabia shift in line with this transition, with longer gaps between occupations phases. However, we must emphasise that direct ages have not been attained for SAHPs 20-17 and 19-13, meaning this argument is currently somewhat tentative. Attaining direct ages for these SAHPs should be a target of future analysis.

The early appearance of *H. heidelbergensis* at Melka Kunture (Ethiopia) soon after ~875 ± 10 ka (MIS 21; Profico et al., 2016) and subsequent appearances in Eurasia is a key event in Early-Middle Pleistocene hominin evolution. SAHPs within this period provide potential timings for *H. heidelbergensis* dispersal, assuming an African origin for this species. While there is a paucity of absolute dating, Oldowan and Acheulean tool typologies have been uncovered in Southern Arabia (Chauhan, 2009; Groucutt and Petraglia, 2012 and references therein; Bailey et al., 2015; Bretzke et al., 2016). This suggests an additional behavioural adaptation was not required for Mode-1 or Mode-

2 bearing hominins to occupy Arabia, at least as represented by the lithic record, but occupation was likely dependent on periods of ameliorated climatic conditions. Here, we have provided timings in which the Arabian Peninsula was occupiable and traversable.

SAHPs during MIS 17 (~700 ka: SAHP 14), MIS 15e (~600 ka: SAHP 13), MIS 15a (~675 ka: SAHP 12), MIS 13c (~530 ka: SAHP 11) and MIS 13a (~480 ka: SAHP 10) may have facilitated dispersals from Africa. Y99 $\delta^{18}O_{ca}$ have positive (SAHP 14 and 13) and variable (SAHP 12, 11 and 10) values, indicating somewhat drier or more variable climates, respectively. Nevertheless, these values are more negative than Holocene values, demonstrating the climate was significantly wetter. As stated above, Oldowan and Acheulean sites are distributed across western and Southern Arabia (Groucutt and Petraglia, 2012), most likely representing Lower Palaeolithic occupation. In particular, flint scatters from the Nafud – found in conjunction with grassland fauna – indicates hominin occupation of Arabia during MIS 13 or 9 (Rosenberg et al., 2013; Stimpson et al., 2016; Roberts et al., 2018). While these SAHPs facilitated occupation of Arabia, it is difficult to relate these to demographic changes in Eurasia due to the persistent presence of Acheulean typologies since ~1.4 Ma (Moncel et al., 2015; Gallotti, 2016). Nonetheless, future studies of Middle Pleistocene population dynamics in Eurasia may benefit from consideration of SAHP timings.

4.3.2 Early H. sapiens *dispersal*

H. sapiens emerged as a distinct species in Africa during the Middle Pleistocene (Hublin et al., 2017; Richter et al., 2017; Scerri et al., 2018b). Behavioural and anatomical modernity evolved gradually throughout the later Middle Pleistocene and into the Upper Pleistocene (Mcbrearty and Brooks, 2000). Whilst *H. sapiens* dispersals during the Upper Pleistocene led to colonisation of Eurasia, it has been suggested that *H. sapiens* may also have dispersed within the Middle Pleistocene (Breeze et al., 2016). This may be validated by a *H. sapiens* maxilla from Misliya Cave, Israel, dated between 194 and 177 ka (Hershkovitz et al., 2018; but see Sharp and Paces, 2018). Hershkovitz et al. (2018) suggested a dispersal may have occurred during MIS 6e (~191-170 ka). Similarly, the recent identification of *H.*

sapiens at Apidima, Greece, ~210 ka (MIS 7) provides further support for earlier dispersals (Harvati et al., 2019; but see Wade, 2019).

Travertine deposition in the Negev (Waldmann et al., 2010), increase rainfall in the Levant and Dead Sea catchment (Frumkin et al., 1999; Bar-Matthews et al., 2003; Gasse et al., 2015; Torfstein et al., 2015), decreased RC09-166 $\delta D_{leafwax}$ (Tierney et al., 2017), and increased Saharan run-off (Williams et al., 2015) indicate ameliorated conditions which may have facilitated *H. sapiens* dispersal within MIS 6e (~191-170 ka) (Breeze et al., 2016; Garcea, 2016). Lack of speleothem growth at Mukalla and Hoti Cave and absence of lake formations (Rosenberg et al., 2011, 2012, 2013) indicate the tropical rain belt did not migrate past 14°N during MIS 6. Moreover, absence of speleothem growth in the central Negev (Vaks et al., 2010) also demonstrates that winter Mediterranean precipitation regimes were not substantially enhanced during MIS 6e. Without these widespread changes in regional precipitation, it is unlikely that green landscapes and inter-regional range expansion could have been sustained.

We therefore suggest dispersals may have occurred during SAHP 5 and 6 (MIS 7a and 7e). Y99 $\delta^{18}O_{ca}$ reveals SAHP 5 (MIS 7a) was as wet as SAHP 4 (MIS 5e: discussed below), and $\delta^{13}C_{ca}$ demonstrate a C_4 biome flourished (Fig. 7). Moreover, SAHP 5 (~205-195 ka) corresponds to lake formations in northern Arabia (Rosenberg et al., 2013), and the Sahara (Armitage et al., 2007, 2015), central ages of palaeosol formation on the Sinai Peninsula (Roskin et al., 2013) and speleothem growth in the central Negev (Vaks et al., 2010). Thus, not only did pluvial landscapes connect northern Arabia and the Levant (Breeze et al., 2016), but corridors connected northern and Southern Arabia. While archaeological evidence for such an early dispersal is currently very limited, recent dating of the Saffaqah archaeological site demonstrates hominins occupied Arabia during MIS 7, with techno-cultural similarities to Mieso (Ethiopia) archaeological assemblages (Scerri et al., 2018a). Further evidence is required, however, taken with recent findings from the Levant (Hershkovitz et al., 2018) and south-eastern Europe (Harvati et at., 2019), our data suggests MIS 7a enhancement of the monsoon domain could have facilitated *H. sapiens* range expansion into Eurasia.

Furthermore, our data shows that Southern Arabia could have facilitated occupation by *H. sapiens* shortly after their African emergence during MIS 9 (Hublin et al., 2017; Richter et al., 2017). Considering recent discussions of the pan-African origin of *H. sapiens* (Scerri et al., 2018b), we suggest that Arabia may have been frequently occupied by various *H. sapiens* lineages. The role of Green Arabia as a habitat for early *H. sapiens* may therefore add to ongoing discussions concerning localized adaptations and genetic flow between subdivided populations (Scerri et al., 2018b). The identification of favourable conditions in MIS 7 and 9 will remain of interest if *H. sapiens* are not identified: i.e. what prevented their expansion into the region at these times?

4.3.3 Late Pleistocene H. sapiens dispersal

The dispersal of *H. sapiens* during the Late Pleistocene is a topic of intense debate, with models changing frequently with new fossil finds. Generally, there is an acceptance that formation of green landscapes in the Saharo-Arabian desert belt facilitated dispersals during MIS 5e (~128-121 ka), MIS 5c (~104-93 ka) and MIS 5a (85-71 ka) (Bae et al., 2017; Groucutt et al., 2018; Rabett, 2018), which is supported by the Mukalla and Hoti Cave records. During the last 130 ka, SAHP 4 (127.8 ± 0.626 to 120.3 ± 0.399 ka) was the most intense pluvial period (Fig. 5), which is in good agreement with lake records in Arabia (Rosenberg et al., 2011, 2012, 2013; Matter et al., 2015). Furthermore, $\delta^{13}C_{ca}$ values demonstrate a grassland environment flourished in the now desert areas of Yemen (Fig 6), which is corroborated by phytolith evidence from Jabal Faya, UAE (Bretzke et al., 2013). SAHP 3 (MIS 5c) and SAHP 2 (MIS 5a) are consistent with intervals of lake formations in Arabia (Petraglia et al., 2012; Rosenberg et al., 2011, 2012, 2013; Parton et al., 2018) during MIS 5c and 5a and It is clear that *H. sapiens* occupied the now desert interior of Arabia within MIS 5a (e.g., Groucutt et al., 2018).

A subsequent dispersal is argued to have taken place during MIS 4-3 (Mellars, 2006, 2013; Shea, 2008; Rohling et al., 2013; Langgut et al., 2018). A growing body of archaeological evidence shows *H. sapiens* were present in Arabia during MIS 3 (Armitage et al., 2011; Delagnes et al., 2012; Jennings et al., 2016), though occupation may have been limited to punctuated pluvial periods (Groucutt and Petraglia, 2012). Whether these could have facilitated more widespread dispersals is a subject of controversial debate (Mellars et al., 2006; Groucutt et al., 2015a; Bae et al., 2017). The palaeoclimatic evidence for increased rainfall during MIS 4/3 is variable between records. It has been argued that punctuated humid intervals in northern Africa and Levant may have facilitated dispersal during MIS 4-3 (Hoffmann et al., 2016; Langgut et al., 2018). However, the lack of speleothem growth in southern Arabia during MIS 4-3 suggest that the ASM was considerably weaker and did not penetrate into the Arabian Peninsula. Our assumption is supported by more positive $\delta D_{\mathrm{leafwax}}$ values in Core RC09-166 from the Gulf of Aden (Fig. 4), indicating lower monsoonal rainfall compared to MIS 5a or the early to middle Holocene in the Horn of Africa and Afar regions. Additional supporting evidence comes from median grain sizes in cores KL 11 (Red Sea) and KL 15 (Gulf of Aden) which are also larger and indicative of more arid climatic conditions during MIS 4-3 (Fig. 4). In contrast, palaeolake formation in the Nafud (northern Saudi Arabia) (Parton et al., 2018) and fluvial activity at Al-Quwaiayh, central Saudi Arabia (McLaren et al., 2009) Yemen (Delagnes et al., 2012), and in Oman (Blechschmidt et al., 2009; Parton et al., 2013, 2015; Hoffmann et al., 2015) have been dated to early MIS 3. Records in Southern Arabia are not corroborated by speleothem growth at Mukalla Cave or Hoti Cave, indicating that the tropical rain-belt was suppressed. This discrepancy may stem from the potential for fluvio-lacustrine and alluvial records to record high intensity, but brief, storm and flooding events (e.g., Rosenberg et al., 2012; Hoffmann et al., 2015); whereas speleothems require prolonged and more substantial changes in regional rainfall. Moreover, brief and intense storms currently occur under modern climates during hot summers, which could mean MIS 3 alluvial records show precipitation regimes were not greatly different than current conditions (Hoffmann et al., 2015). Taken together, MIS 3 precipitation may not have been sustained or intense enough to have substantially impacted environments in Southern Arabia. Despite archaeological evidence for MIS 3 occupation, our findings suggest that MIS 5 interstadials were ‘climatic optima’ for hominin dispersal; whereas *H. sapiens* dispersal opportunities during MIS 3 may have been more limited or required different behavioral adaptations. If, however, MIS 3 environments could not sustain dispersal (meaning MIS 3 populations can be related to groups that entered during MIS 5), this implies *H. sapiens* persistence during MIS 4, a time with very limited evidence for ameliorated conditions (Parton et al., 2015).

5. Conclusion

New ^{230}Th and U-Pb dates for stalagmite Y99 from Mukalla Cave in Yemen allow to extend the speleothem-based record of continental wetness back to 1.1 Ma BP. In combination with previously published stalagmite records from Southern Arabia (Burns et al., 2001; Fleitmann et al., 2011), at least twenty-one humid intervals with precipitation above ~300 mm yr^{-1} developed in Southern Arabia, all them occurred during peak interglacial periods. Of all SAHPs the early to mid-Holocene humid period was the least humid period. Hydrogen and oxygen isotope measurements on water extracted from stalagmite fluid inclusions indicate that enhanced rainfall during SAHPs resulted from an intensification and greater range of the ASM and ISM. This assumption is further supported by the presence of annual laminae in some stalagmites and nearly monthly-resolved oxygen and carbon isotope measurements which indicate a strong seasonal climate during SAHP, with one rainy (monsoon) season during SAHPs. While there is restricted archaeological evidence for hominin occupation beyond 350 ka BP, these landscapes *could* have facilitated occupation of late *H. erectus* populations. Subsequent SAHPs may have also facilitated the dispersals of early *H. sapiens* soon after their emergence ~300 ka BP.

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Appendix A. Supplementary data

The following are the supplementary data to this article:

Fig. S1:

Fig. S2:

Fig. S3:

Tab. S1-S3:

Tab. S4-S5:

Tab. S5:

Tab. S6:

Tab. S7:

Tab. S8:

Tab. S9:

Tab. S10:

Tab. S11:

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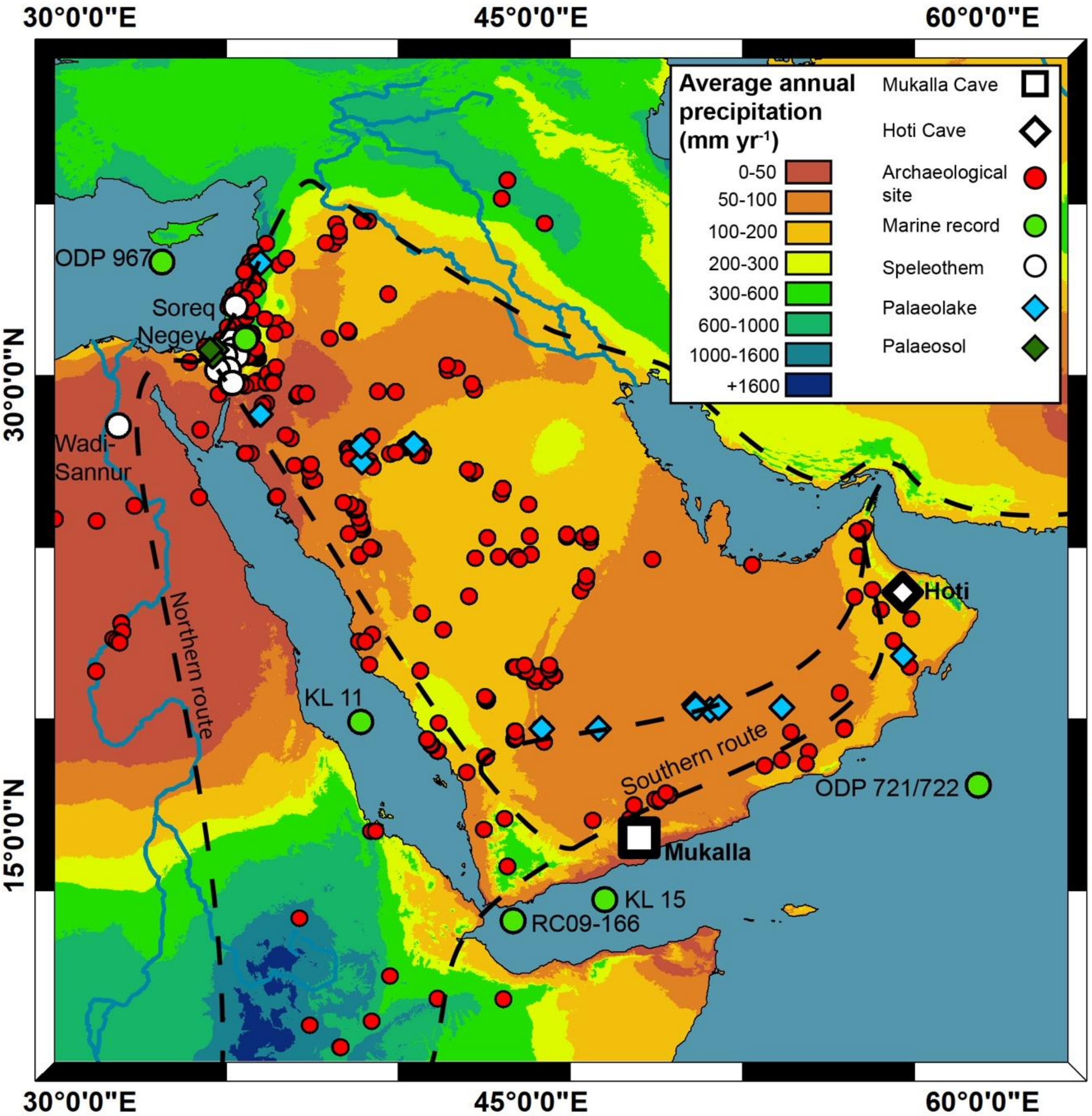


Fig. 1. Map of the Arabian Peninsula with present day (1970-2000) annual precipitation (accessible at: worldclim.org*; Fick and Hijmans, 2017). Red circles denote Middle Palaeolithic archaeological sites (red circles; Bailey et al., 2015; Breeze et al., 2015; Groucutt et al., 2015a, 2015b; Jennings et al., 2015a; Petraglia et al., 2011; Rose et al., 2011). Dashed lines show potential hominin dispersal routes (Rosenberg et al., 2011). Also shown are caves (white cirlces; Bar-Matthews et al., 2003; El-Shenawy et al., 2018; Frumkin et al., 1999; Vaks et al., 2010); palaeolake sites (blue diamonds; Rosenberg et al., 2011, 2012, 2013; Petraglia et al., 2012; Matter et al., 2015), marine records (green circles; deMenocal, 1995; Fleitmann, 1997; Almogi-Labin et al., 2000; Larrasoana et al., 2003; Tierney et al., 2017), lake*

records (blue circles; Torfstein et al., 2015) and Mukalla and Hoti caves (hollow square and diamond, respectively; Burns et al., 1998, 2001; Fleitmann et al., 2003b, 2011).

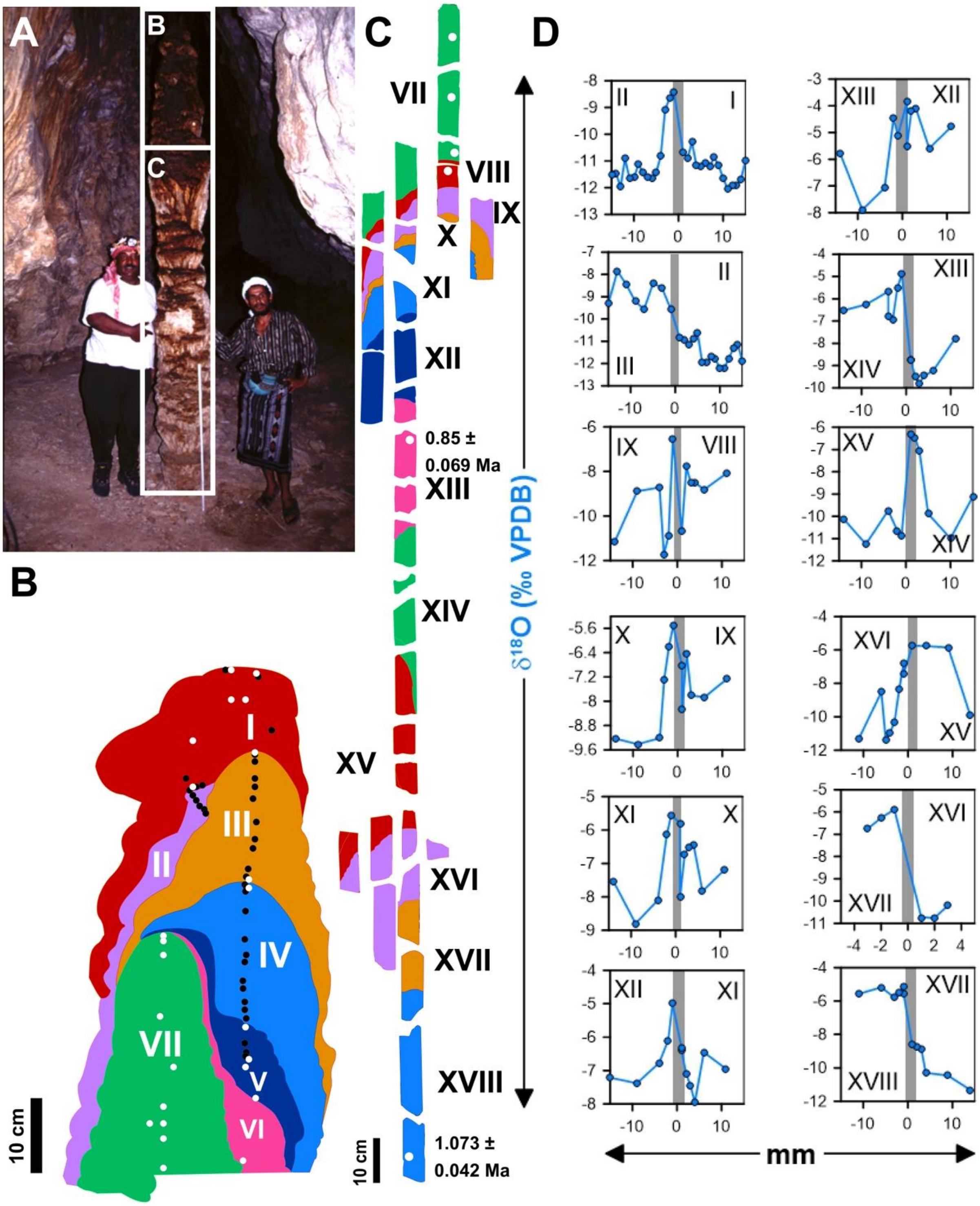


Fig. 2. (A) Stalagmite Y99 in situ in Mukalla Cave. (B and C) Y99 consecutive growth intervals (Fig. S1-S3). Location of ^{230}Th and U-Pb ages marked by black (Fleitmann et al., 2011) and white (this study) circles. (D) Plots show $\delta^{18}O_{ca}$ shifts over discontinuities between GIs (Tab. S6 and S7).

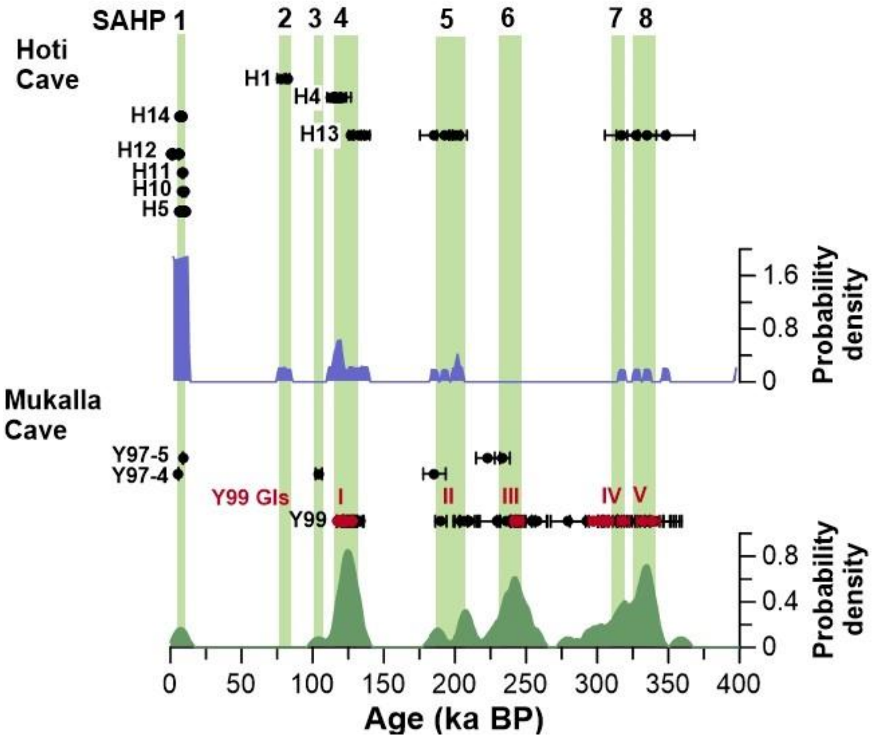
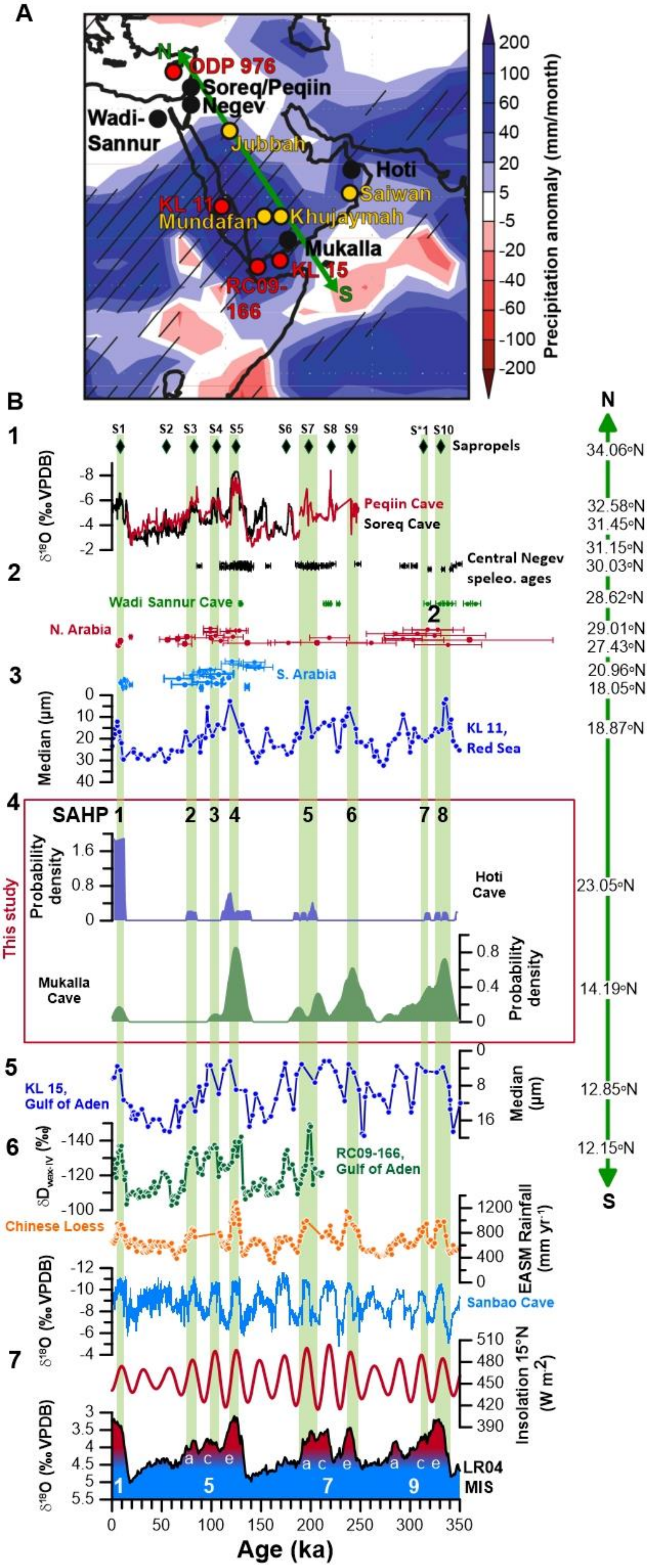


Fig. 3. ^{230}Th ages of Hoti Cave and Mukalla Cave speleothems. Red dots denote new Y99 ^{230}Th ages determined for this study. Age kernel probability density plots of Hoti (blue; 5 pt. moving average) and Mukalla (green) and green bars show periods of most likely speleothem deposition. These were used to assign South Arabian Humid Periods (SAHP) 1-8.



A
N
ODP 976
Soreq/Peqiin
Negev
Wadi-
Sannur
Jubbah
Hoti
Saiwan
KL 11
Mundafan
Khujaymah
Mukalla
RC09-
166
KL 15
S
Precipitation anomaly (mm/month)
B
Sapropels
Peqiin Cave
Soreq Cave
Central Negev speleo. ages
Wadi Sannur Cave
N. Arabia
S. Arabia
KL 11, Red Sea
SAHP
This study
Hoti Cave
Mukalla Cave
Probability density
KL 15, Gulf of Aden
RC09-166, Gulf of Aden
Chinese Loess
EASM Rainfall (mm yr-1)
Sanbao Cave
Insolation 15°N (W m-2)
LR04
MIS
Age (ka)

Fig. 4. (A) Location of speleothems (black circles), palaeolakes (yellow circles) and marine sediment cores from the eastern Saharo-Arabian deserts compared to simulated precipitation anomalies (MIS 5e – pre-industrial: Herold and Lohmann, 2009). (1) Sapropel layers in the Eastern Mediterranean Sea (green diamonds; Williams et al., 2015; Grant et al., 2017) vs Peqiin Cave (red) and Soreq Cave (black: Bar-Matthews et al., 2003) $\delta^{18}O$ records; (2) Central Negev (black; Vaks et al. 2010) and Wadi Sannur (green; El-Shenawy et al. 2018) speleothem deposition periods compared to north (red; Petraglia et al., 2012; Rosenberg et al., 2013; Parton et al., 2018) and South Arabian lakes (blue; Rosenberg et al., 2011, 2012; Matter et al., 2015); (3) Red Sea median grain size (Fleitmann, 1997); (4) age kernel density plots of Hoti Cave (blue; 5pt moving average) and Mukalla Cave (green) stalagmites; (5) Gulf of Aden median grain size (Fleitmann, 1997) and $\delta D_{leafwax}$ ‰ (Tierney et al., 2017); (6) Chinese reconstructed rainfall (Beck et al., 2018) vs Sanbao Cave composite speleothem $\delta^{18}O_{ca}$ record (Cheng et al., 2016); (7) Northern hemisphere June insolation at 15°N (Berger and Loutre, 1991, 1999) vs global marine $\delta^{18}O_{benthic}$ (Lisiecki and Raymo, 2005). Marine Isotope Stages follow the taxonomy of Railsback et al. (2015).

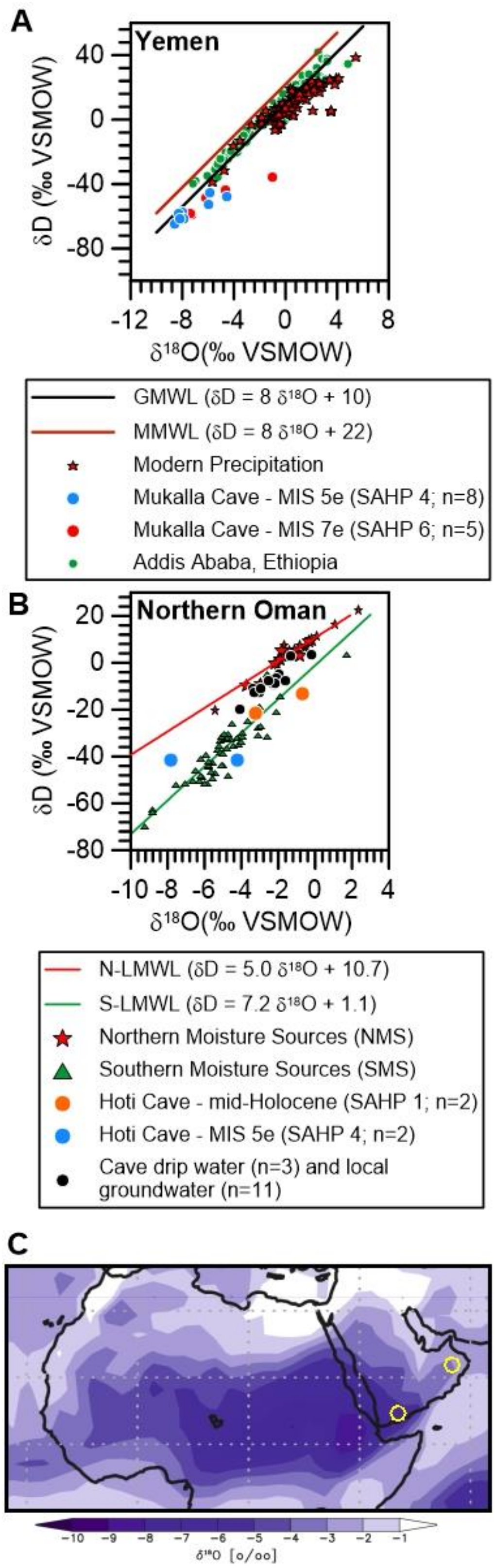


Fig. 5. Water isotope (δD_{FI} and $\delta^{18}O_{FI}$) values from stalagmites from Mukalla and Hoti Caves (Tab. S9).

(A) Stalagmite Y99 δD_{FI} and $\delta^{18}O_{FI}$ values in comparison to δD and $\delta^{18}O$ in modern precipitation in Yemen (Al-ameri et al., 2014) and Ethiopia (IAEA/WMO, 2019. Global Network of Isotopes in

Precipitation. The GNIP Database. Accessible at: https://nucleus.iaea.org/wiser). Black line denotes the Global Meteoric Waterline (G-MWL: δD = 8 δ^{18}O + 10). Brown line denotes the Mediterranean Meteoric Waterline (δD = 8 δ^{18}O + 22) (Gat and Carmi, 1970; Matthews et al., 2000; McGarry et al., 2004). (B) δD_{FI} and $\delta^{18}O_{FI}$ values H5 and H13 compared to regional precipitation values and meteoric waterlines from Northern Oman (N-LMWL: δD = 5.0 δ^{18}O +10.7; Weyhenmeyer et al., 2000, 2002) and Southern Oman (S-LMWL: δD = 7.2 δ^{18}O +1.1; Weyhenmeyer et al., 2000, 2002). (C) Locations of Mukalla Cave and Hoti Cave relative to modelled $\delta^{18}O_{precipitation}$ values for boreal summer precipitation during MIS 5e (modified after Herold and Lohmann, 2009). Yellow circles mark the location of Mukalla and Hoti Caves.

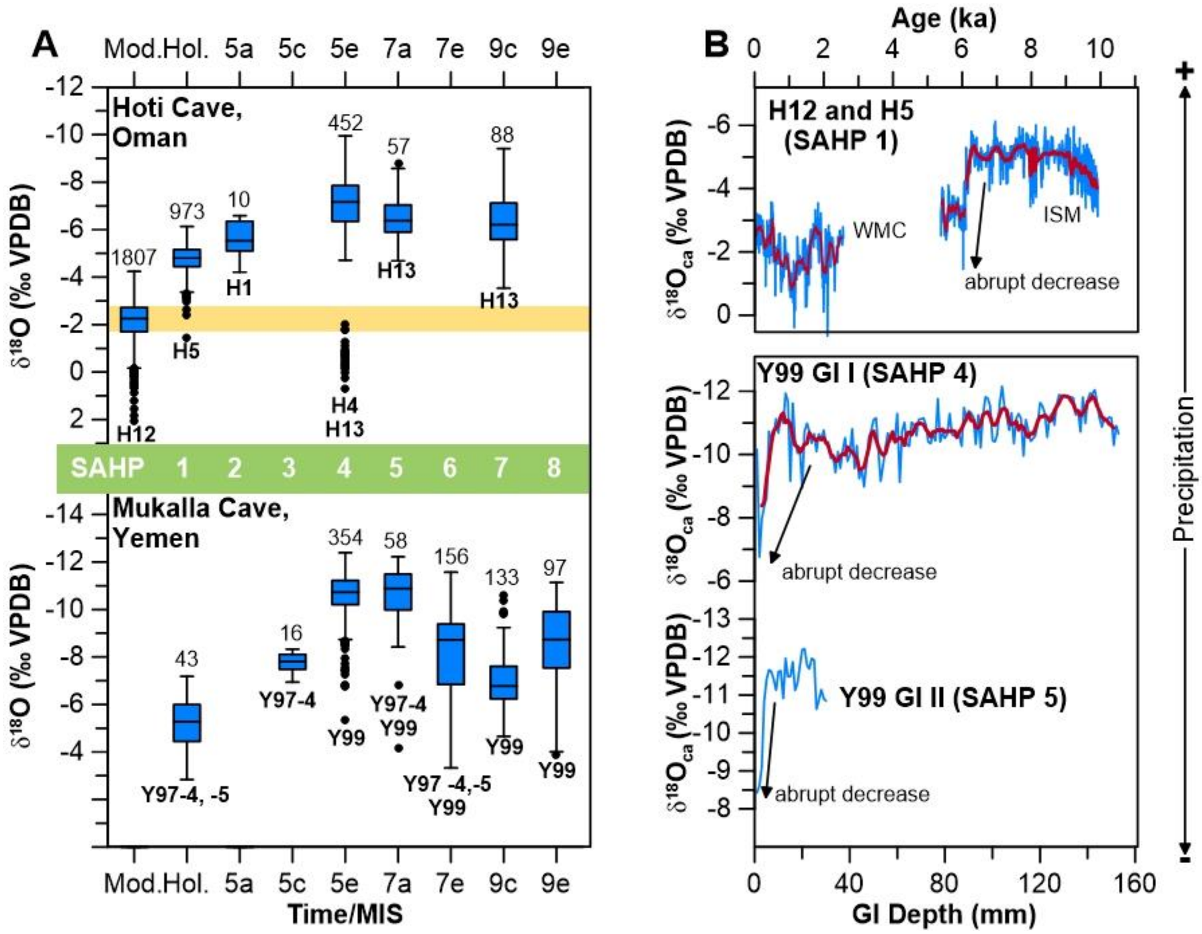


Fig. 6. (A) $\delta^{18}O_{ca}$ whisker-boxplot of Mukalla Cave and Hoti Cave composite records (new Y99 $\delta^{18}O_{ca}$ values combined with data from Fleitmann et al. 2011; Tab. S10). Numbers below whiskers denote

sample labels and number of $\delta^{18}O_{ca}$ measurements Statistically extreme values marked as black circles. (B) $\delta^{18}O_{ca}$ profiles of Holocene (H5 and H12) and MIS 5e (Y99 GI I) and MIS 7a (Y99 GI II) stalagmites.

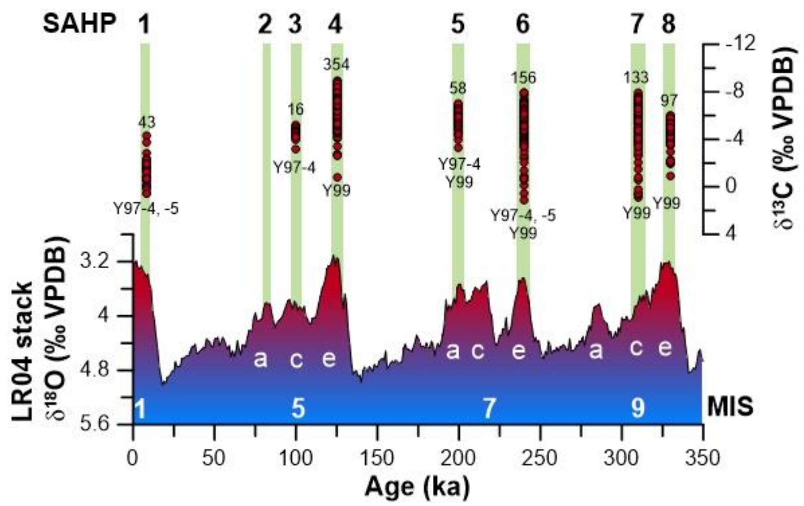
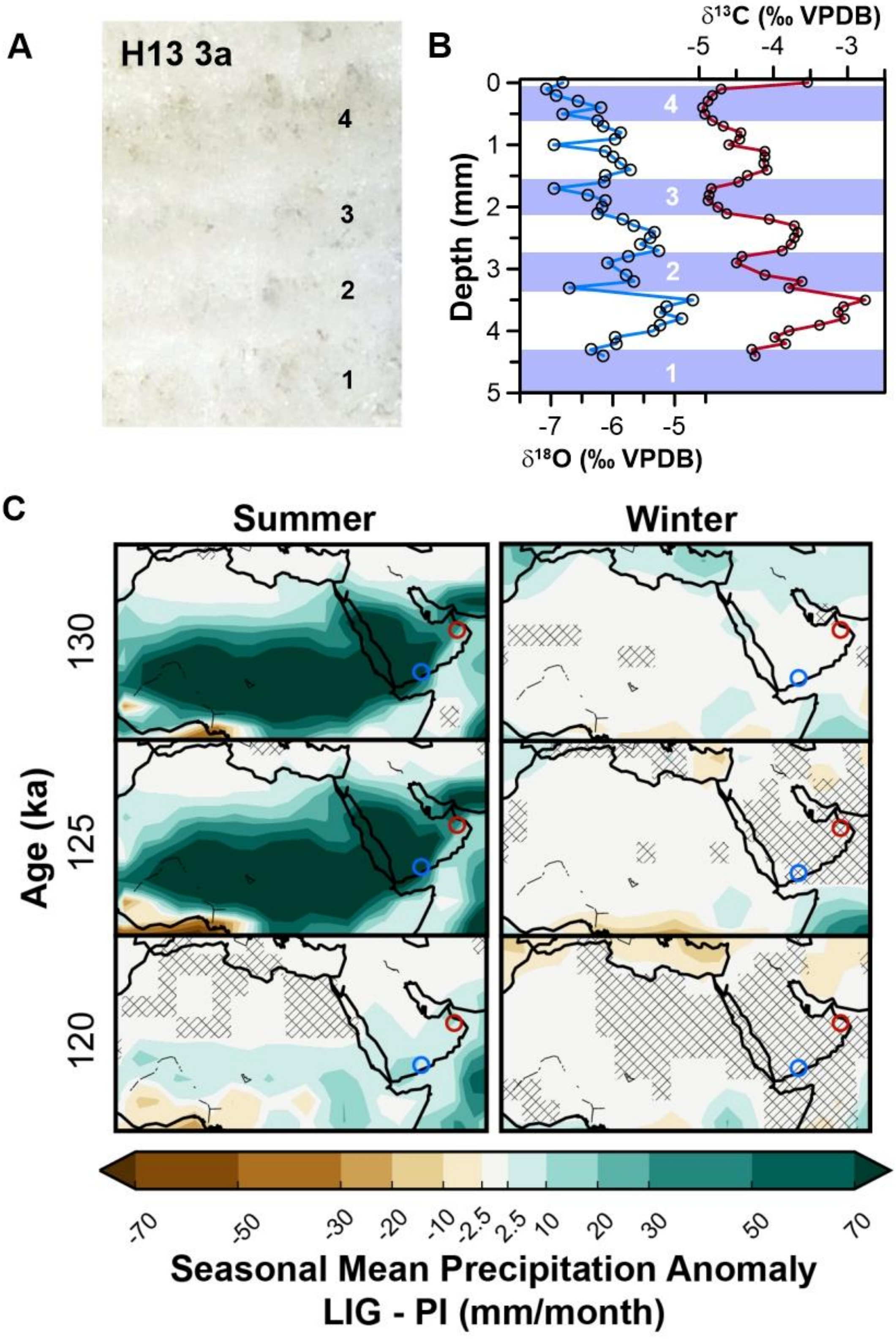


Fig. 7. $\delta^{13}C_{ca}$ values of Mukalla Cave speleothems during SAHPs I-V (Tab. S11) compared to the LR04 stack $\delta^{18}O$ record (Lisiecki and Raymo, 2005).



A
H13 3a
4
3
2
1
B
δ13C (‰ VPDB)
-5
-4
-3
Depth (mm)
0
1
2
3
4
5
-7
-6
-5
δ18O (‰ VPDB)
C
Summer
Winter
Age (ka)
130
125
120
-70
-50
-30
-20
-10
-2.5
2.5
10
20
30
50
70
Seasonal Mean Precipitation Anomaly
LIG - PI (mm/month)

1177

Fig. 8. (B) Sub-annual $\delta^{18}O_{ca}$ and $\delta^{13}C_{ca}$ values from a MIS 5e section of stalagmite H13 (A) from Hoti Cave (Tab. S8). Shaded blue areas and numbers mark the monsoon seasons. (C) Mukalla Cave (blue circle) and Hoti Cave (red circle) mapped to modelled MIS 5e winter and summer precipitation anomaly (compared to pre-industrial) (Gierz et al., 2017).

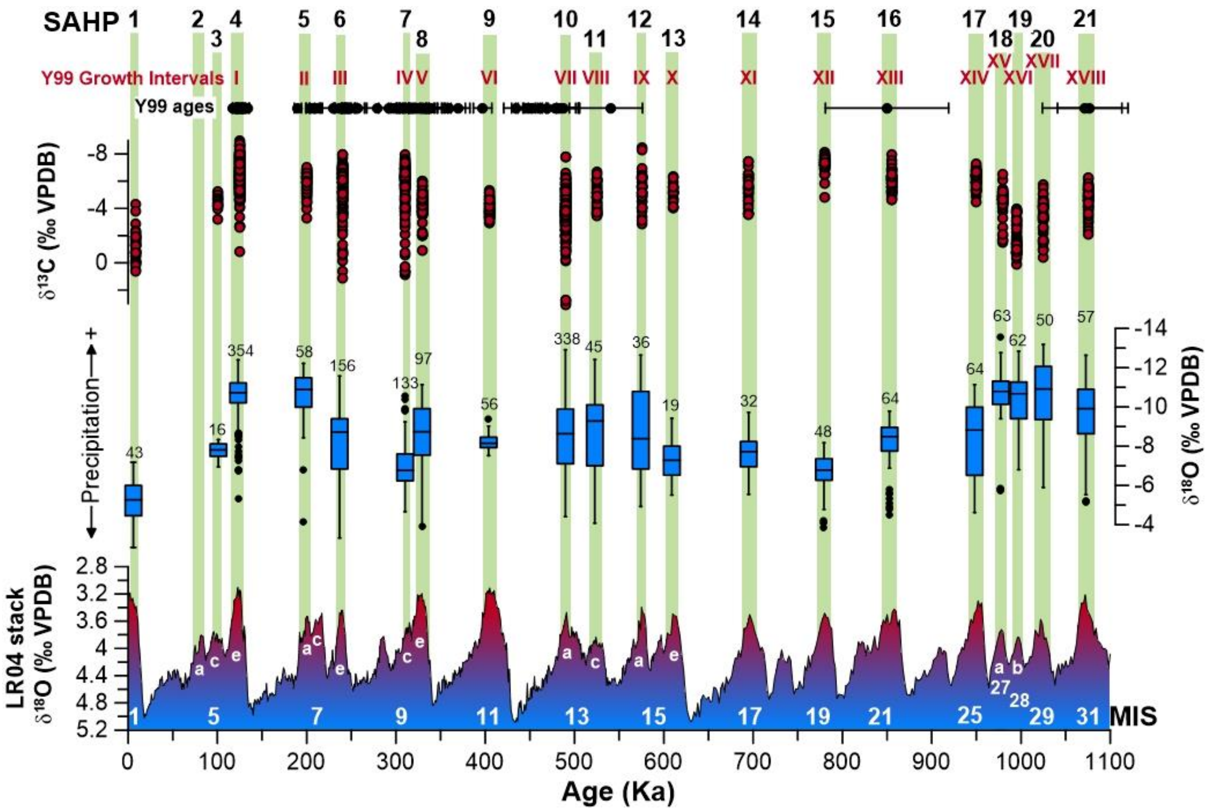


Fig. 9. ^{230}Th (Tab. S1-S3) and U-Pb (Tab. S4 and S5) ages for stalagmite Y99 compared to the LR04 stack $\delta^{18}O$ record (Lisiecki and Raymo, 2005) and extended $\delta^{18}O_{ca}$ and $\delta^{13}C_{ca}$ records of Mukalla Cave stalagmites (Y97-4, Y97-5 and Y99). Undated Y99 growth intervals were assigned to intermediate interglacials and warm substages. Green bars denote timing of SAHPs (South Arabian Humid Periods). Marine Isotope Stages follow the taxonomy of Railsback et al. (2015).

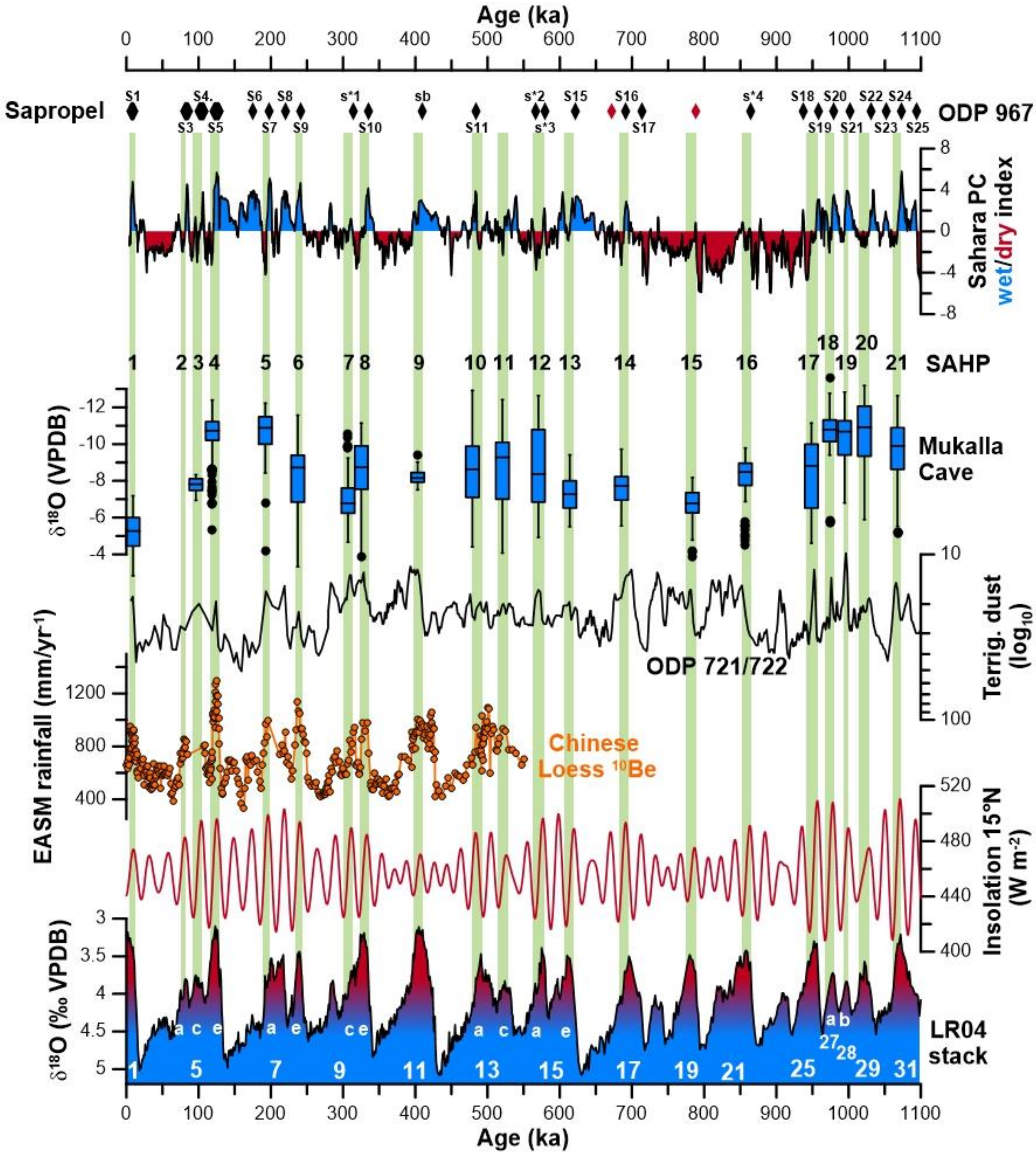


Fig. 10. SAHPs (green bars) and palaeoclimate records. (Eastern Mediterranean) ODP 967 sapropels (black = identified, red = 'ghost') and wet/dry PCA model (Grant et al., 2017); central Negev desert speleothem ages (Vaks et al., 2010); northern and Southern Arabian palaeolake ages (Rosenberg et al., 2011, 2012, 2013; Matter et al., 2015; Parton et al., 2018) and Mukalla Cave $\delta^{18}O_{ca}$ values; ODP 721/722 terrigenous dust (deMenocal, 1995); EASM reconstructed rainfall from Chinese $^{10}Be_{loess}$ (Beck et al., 2018); NHI insolation (W m^{-2}) at 15°N (Berger and Loutre, 1991, 1999) and LR04 stack

foraminifera $\delta^{18}O_{benthic}$ (Lisiecki and Raymo, 2005) and Marine Isotope Stages following the taxonomy of Railsback et al. (2015).